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Soil Carbon Dynamics

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The amount of organic carbon (C) stored in the upper meter of mineral soils in the Amazon Basin (~40 Pg C) represents ~3% of the estimated global store of soil carbon. Adding surface detrital C stocks and soil carbon deeper than 1 m can as much as quadruple this estimate. The potential for Amazon soil carbon to respond to changes in land use, climate, or atmospheric composition depends on the form and dynamics of soil carbon. Much (~30% in the top ~10 cm but >85% in soils to 1 m depth) of the carbon in mineral soils of the Oxisols and Ultisols that are the predominant soil types in the Amazon Basin is in forms that are strongly stabilized, with mean ages of centuries to thousands of years. Measurable changes in soil C stocks that accompany land use/land cover change occur in the upper meter of soil, although the presence of deep roots in forests systems drives an active C cycle at depths >1 m. Credible estimates of the potential for changes in Amazon soil C stocks with future land use and climate change are much smaller than predictions of aboveground biomass change. Soil organic matter influences fertility and other key soil properties, and thus is important independent of its role in the global C cycle. Most work on C dynamics is limited to upland soils, and more is needed to investigate C dynamics in poorly drained soils. Work is also needed to relate cycles of C with water, N, P, and other elements.

1. INTRODUCTION: WHY IS SOIL CARBON IMPORTANT?

Globally, soils store at least twice as much carbon (C) as the atmosphere. Hence, changes in soil C stores can potentially play an important role in interannual to decadal variations in the global C cycle, and management of C during land use change could be significant in terms of regional or national C commitments. In the Amazon Basin, uncertainties about the degree to which stores of soil C in the large areas of intact forest will change with climate and elevated CO_2 or with future land cover or land use change are part of what limits our ability to estimate future feedbacks between ecosystem C stocks and atmospheric CO_2 levels [*Cox et al*., 2000; *Friedlingstein et al*., 2006]. Intact Amazon tropical forests have been suggested to be current carbon sinks to a degree that nearly offsets deforestation losses [*Stephens et al.*, 2007], though the role of soils is uncertain in this balance. Some models predict large soil C losses from intact forests under scenarios of future climate change [*Falloon et*

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al., 2007], and land clearing also has the potential to destabilize soil C. It is, thus, necessary to examine the capacity for Amazon soils to store or lose carbon, which, in turn, requires knowing (1) how much C is stored in soils and (2) how rapidly it can be stabilized or destabilized.

Soil organic matter controls key soil properties, in particular, those associated with overall nutrient availability, water holding capacity, and fertility that make it important beyond its role in the global C cycle. Decomposition processes in soils are major sources or sinks of important non-CO_2 trace gases (methane, oxides of nitrogen, carbon monoxide). Further, the presence of Amazon "black earth" soils, anthropogenically altered soils with large stocks of stabilized organic matter and high fertility, suggest the potential to manage even soils generally considered unfertile to store carbon [*Woods*, 2003].

This chapter will focus on current understanding of the dynamics of carbon in Amazonian soil organic matter and comment on the potential role of Amazon soils in the perturbed global carbon cycle. It will summarize the key factors needed to determine soil C dynamics: (1) the total inventory of organic C in soils; (2) the fluxes of C into and out of soils; and (3) the age of C stored in and respired from soil. The potential for soil organic matter to behave as a source or sink of CO_2 can then be assessed from estimates of how C fluxes into and out to the soil pool are altered by climate, vegetation, or land use change and the time C resides in soil organic matter. Each of these soil properties, in turn, is related to soil forming factors: climate, organisms (vegetation and soil fauna), parent material, topographic position, and time [*Jenny*, 1947].

Studies of Amazon soil carbon and its dynamics have tended to focus on old, highly weathered soils developed under terra firme tropical forest or *cerrado* vegetation that cover most of the area of the Basin. Much of the attention of these studies has been on the influence of land use and its effects on C and nutrient cycling and ultimately soil fertility [see *Tiessen et al.*, 1994; *Lehmann et al.*, 2001; *Cerri et al.* 2007a, 2007b; *Paul et al.*, 2008]. In contrast, this review will focus on the dynamics and cycling of carbon in intact forest soils and their role in the overall ecosystem carbon cycle. It will also focus on upland, well-drained soils; further studies are needed to understand C dynamics in seasonally flooded and wetland soils.

2. ORGANIC CARBON STOCKS IN AMAZON SOILS

Most estimates of the amount of carbon stored as organic matter in Amazonian soils are based on profile data collected by the Radambrasil Project (1973–1986) and compiled in the SOTER database [*Batjes*, 2005; *Batjes and Dijkshoorn*, 1999]. Researchers attempting to make basin-wide estimates of C storage from these profile data have used different assumptions of how to fill in missing data for bulk density, depths between samples, and to extrapolate spatially from limited profile data; summaries can be found in Table 1 of *Batjes* [2005] or Table 2 of *Cerri et al*. [2007b]. Estimates of Amazonian soil C stocks to 1 m depth range from 41 to 47 Pg C for the ~500 Mha of the Brazilian Amazon [*Cerri et al*., 2007b; *Moraes et al.*, 1995]; *Batjes and Dijkshoorn* [1999] report a total storage of 66.9 Pg C for the top 1 m of the total ~700 Mha of the Amazon Basin. Some 44–67% of the C stored in the top meter is present in the 0–30 cm depth interval [*Bernoux et al*., 2006].

Most (50–75%) of the area of the Amazon basin is represented by two soil orders: Acrisols (Ultisols, US Taxonomy or Podzólico, Brazilian taxonomy) and Ferralsols (Oxisols or Latossolo); other important soil orders include Gleysols and Leptosols (each about 8% of the area) [*Batjes and Dickshorn*, 1999; *Cerri et al*., 2000, 2007b]. C stocks in Oxisols and Ultisols soils are on average similar, from 44 Mg C ha^{-1} (Ultisols) to 55 Mg C ha^{-1} (Oxisols) in the top 30 cm and 85 (Ultisols) to 100 Mg C ha^{-1} (Oxisols) in the top 100 cm, respectively [*Batjes and Dijkshorn*, 1999]. Carbon to nitrogen ratios for 0–30 cm are ~10 (Ultisols) to 13 (Oxisols), decreasing to values of 9–11 by ~1 m depth [*Batjes and Dijkshorn*, 1999].

Soils within the same Order tend to have lower C stocks in Acre State (western Amazon) than their counterparts in the central and eastern Amazon [*Melo*, 2003]. For example, average C stocks in Acre state are estimated at 38 (0–30 cm) and 66 Mg C ha^{-1} (0–100 cm) [*Melo*, 2003]. In general, dystrophic soils (more highly weathered, with lower base saturation and lower fertility) tend to have higher C stocks than eutrophic soils (less weathered, higher base saturation, and more fertile soils). Soils in the western Amazon tend to be younger and more eutrophic, which may explain lower C stocks observed by *Melo* [2003].

Soil C stocks vary locally with factors like topography and land management, and more regionally with soil parent material and underlying geology [*Cerri et al*., 2004; *Holmes et al*., 2004, 2006]. In local studies of soil C and N stocks along topographic gradients near Manaus and Santarem, soils graded from Oxisols on plateaus, to Ultisols on slopes, and Spodosols in valleys (Arenosols in the Brazilian classification) [*Luizão et al*., 2004; *Telles et al.*, 2003]. Spodosols in periodically flooded lowlands that dissect plateaus at sites near Manaus and Santarem have distinctly higher sand and lower clay contents, with lower soil C stocks and higher C:N ratios in the upper 10–40 cm of soil [*Luizão et al*., 2004; *Telles et al.*, 2003].

The published estimates for soil C stocks do not include superficial litter [*Batjes and Dijkshorn*, 1999], which can add

a substantial amount of rapidly cycling carbon to these totals (Table 1). Leaf litter inventories vary seasonally but contain from 3 to 7 Mg C ha^{-1} in primary forests [*Dantas and Phillipson*, 1989]. Stocks of litter can be larger in secondary growth forests and are generally much smaller in agroforests and pastures. Estimates of coarse (>10 cm diameter) woody debris on in mature forest floor range from ~10 to 35 Mg C m^{-2} (assuming biomass is 50% C) [*Chambers et al.*, 2001b; *Keller et al.*, 2004; *Rice et al.*, 2004]. Other litter components, branches, fruits, etc., can make up an additional 15 Mg C ha^{-1} [*Chambers et al.*, 2001b], though this component is not often quantified. Hence, detritus (also referred to as necromass) on the forest floor contains roughly the same amount of C as is found in the upper 30 cm of mineral soil.

Soil carbon stocks calculated to 1 m depth may be considered a minimum for total soil C storage, as many Amazonian forests are deep-rooting [*Nepstad et al.*, 1994]. Even though the concentrations of C in soils decrease dramatically with depth, the large volumes of soil mean that the total amount of C stored from 1 to 8 m depth can equal what is stored in the top 30 cm [*Trumbore et al*., 1995; *Camargo et al.*, 1999]. Root standing stocks belowground are also seldom considered in estimates of soil carbon. *Telles et al*. [2003] found that fine root biomass was 2–2% of total C stocks in the upper meter of primary forest soils (after sieving to remove objects >2 mm), and roughly a third of that in the 0–10 cm layer. Live and dead fine roots clearly drive active C and water cycles deeper than 1 m [*Nepstad et al.*, 1994; *Oliveira et al.*, 2005; *Trumbore et al.*, 2006; *Fisher et al.*, 2007].

Surface soil (0–20 or 0–30 cm) C stores in Amazon soils are generally low compared to temperate or boreal soils [*Sanchez and Buol*, 1975]. The 40–50 Pg C in the top meter of Brazilian Amazon soils (66 Pg C for the whole basin) represents only about 2–5% of global mineral soil C stocks to 1 m depth, though the Amazon represents roughly 14% of land area. Including necromass (surface detritus) and soil depths greater than 1 m more than double estimates of total C stored in intact forest soils of the Amazon (Table 1), making soil C stocks roughly equivalent to aboveground carbon stocks for mature forests. The question of most interest for understanding the role of Amazonian soil C in the global carbon budget, however, is not how much C is stored, but what fraction of that organic carbon is in forms that can accumulate or be released on timescales of the next decades to centuries.

Table 1. Representative Carbon Stocks in Unmanaged Terra Firme Forest (Oxisols)

Stock	Inventory (Mg C ha^{-1})	Reference
Surface litter		
Fine	3–7	*Dantas and Phillipson* [1989], *Selva et al.*, [2007]
Medium	15	*Chambers et al.* [2001b]
Coarse	10–35	*Chambers et al.* [2001b], *Keller et al.* [2004], *Rice et al.* [2004]
Root biomass[a]		
0–10 cm	0.2–0.5, *0.6–1.2*	*Silver et al.* [2004]
10–100 cm	0.5, *0.7*	*Trumbore et al.* [2006]
100–800 cm	0.6, *0.7*	
Soil organic matter		
0–30 cm	44–55	*Batjes and Dijkshorn* [1999]
30–100 cm	28–47	
100–300 cm	~70	*Camargo et al.* [1999]
300–800 cm	~100	

[a]Live biomass is shown in roman; dead biomass is shown in italic.

3. CARBON FLUXES INTO AND OUT OF AMAZON FOREST SOILS

The C added annually to soils includes fine and coarse litterfall, tree mortality, and root mortality. Only a small fraction of this added carbon ends up stored as soil organic matter; the vast majority is decomposed to CO_2 [*Luizão and Schubart*, 1987; *Parton et al.*, 2007]. Relatively few measures of C inputs are available in the literature for Amazonian forests (Table 2a), and this is a major limitation to understanding forest C dynamics and how it varies in space and time. Surface litter fluxes range from 2 to 4 Mg C ha^{-1} a^{-1} (summarized by *Cattiano et al.* [2004]), some 40–67% by mass of which is leaves [*Martius et al.*, 2004; *Selva et al.*, 2007]. Fewer estimates of belowground productivity exist; recent estimates are ~1 Mg C ha^{-1} a^{-1} [*Silver et al.*, 2005; *Trumbore et al.*, 2006]. *Vieira et al.* [2005] used dendrometer measurements in the western, central, and eastern Amazon to estimate that C allocated to woody stem growth is ~2 Mg C ha^{-1} a^{-1}. Assuming a steady state forest, this provides an average estimate for woody debris inputs. For the Tapajos forest, stem growth increment is in accord with the estimates of coarse (>10 cm diameter) necromass production by *Palace et al.* [2008]. *Palace et al*. [2008] also estimate medium (5–10 cm diameter) and fine (2–5 cm diameter) branch and woody necromass inputs in an intact forest as adding an additional ~1 Mg C ha^{-1} a^{-1}. Allocation of net primary production (NPP) by the estimates in Table 2a (leaf litterfall: stem/branch growth:root production of ~3–4:2–3:1) contrast with estimates built in to ecosystem models of roughly equal C

Table 2a. Representative C Fluxes in Unmanaged Terra Firme Forests

Flux	Mg C (ha^{-1} a^{-1})	Reference
Litterfall		
<2 cm	3.0–6.2	*Cattiano et al*. [2004]
2–10 cm	1.0	*Palace et al.* [2008]
>10 cm	2.3	*Palace et al.* [2008]
	2.0–2.2[a]	*Vieira et al.* [2005]
Fine root production	0.65–1.0	*Silver et al.* [2005], *Trumbore et al.* [2006]
Export as POC	0.0176	*Johnson et al.* [2006], *Selva et al.* [2007]
Export as DOC	0.0315	*Johnson et al.* [2006]

[a]More than 10 cm inputs derived from estimates of annual woody biomass increment and the assumption of steady state.

Table 2b. Fluxes of Carbon in Soil Respiration, Litterfall, and Total Belowground Carbon Allocation in Selected LBA Sites[a]

	Soil Respiration	Litter-fall	TBCA[b]	Reference
Paragominas (PA)	20.0	4.3	15.7	*Davidson et al.* [2000]
Tapajos Forest (PA)	12.8	5.7–6.3	6.5–7.1	*Brando et al.* [2008], *Rice et al*. [2004]
Manaus (AM)	12.1	3.3	8.8	*Chambers et al.* [2004]
Acre (AC)	17.0	4–6[c]	11–13	*Salimon et al.* [2004]

[a]Fluxes are in Mg C ha^{-1} a^{-1}.

[b]TBCA, total belowground C allocation equal to soil respiration minus litterfall.

[c]Litterfall estimated from older secondary forests in Acre.

allocation of NPP to leaf litter, stem growth, and root productivity [e.g., *Fung et al.*, 1997].

Soil respiration rates (Table 2b) integrate CO_2 derived from plant as well as microbial respiration sources and, hence, are larger than estimated soil C inputs. A proposed measure of total belowground C allocation (TBCA) [*Ryan*, 1991] is determined as soil respiration rates minus surface (leaf) litterfall. Using reported soil respiration and litterfall estimates from several studies, greater belowground C allocation occurs in sites with extended seasonal drought (Paragominas and Acre) compared to the central Amazon (Manaus), although the data for the Tapajós forest near Santarem does not follow this pattern (see Table 2b). Given low root productivity (Table 2a), high TBCA values also indicate that the large amounts of C allocated belowground are allocated to forms other than root growth, e.g., root respiration or transfer to the soil through symbiotic fungi or root exudates.

Export of carbon from soils in dissolved inorganic (DIC) or organic forms (DOC) is an important source of C to streams, but these fluxes are small compared to C inputs or soil respiration rates. *Richey et al*. [2002] estimated that most of the C respired by the Amazon River system originates in soils and that this export could be large enough to explain the difference between C uptake reported for eddy covariance studies and what was observed accumulating in vegetation and soils in the footprints of the flux towers. However, estimates of fluxes from upland forests are small (~0.03% of litterfall) and suggest upland litter export does not supply large amounts of particulate C to streams compared to riparian and seasonally flooded areas (Table 2a) [*Selva et al*., 2007]. Soil pore space CO_2 is likely the largest contributor to stream DIC [*Johnson et al.*, 2006; *Selva et al*., 2007]. Dissolved organic C fluxes from upland soils are smaller than DIC fluxes. *Remington et al.* [2007] demonstrated lower sorption capacity of sandier lowland soils compared to upland clay-rich Oxisol soil material, supporting the idea that much of the in-stream source of DOC and even DIC may be in the riparian zone [*McClain et al.*, 1997]. In terms of the net C balance of upland soils, DIC and DOC fluxes are small compared to other terms in the annual soil C budget.

4. DYNAMICS OF SOIL C IN INTACT FORESTS

Carbon in soils is not homogeneous; it consists of a wide range of chemical compounds that differ in their intrinsic rates of decomposition and the degree to which they may be stabilized through interactions with mineral surfaces or inclusion in aggregates. Soil biogeochemistry models like Century or RothC parameterize this by identifying pools of carbon, metabolic and resistant plant residues, microbial biomass, and "active," "slow," and "passive" pools, with decomposition rates that vary from years or less for "active" pools to millennia for "passive" pools. A major and continuing challenge has been to estimate the amount of C in each of these pools from observable characteristics of the soils themselves. Therefore, predicting the response of soil C stocks to changes in land cover or climate requires an understanding of the rates at which stored C is replenished, and the rates at which C may be rapidly stabilized or destabilized when conditions change.

Several methods have been employed for determining the rate at which C cycles through terrestrial ecosystems. The first of these compares the stocks of C stored in soil organic matter with the rate at which new C is added to soils as detrital material or lost by decomposition or leaching. This approach, if applied uniformly across all soil organic matter types, offers a long-term average that likely overestimates the short-term response of soil C to a change [*Trumbore*,

2000]. For example, an estimate based on the C inventories in Table 1 and inputs in Table 2a will vary depending on what soil depth (0–30, 0–100, 0–800 cm) is considered. A second approach is to observe changes in C stocks after some disturbance, though this depends on the availability of appropriate chronosequences for the study of dynamics of C on longer timescales. Carbon isotopes, including ^{13}C as well as cosmogenic and bomb-produced radiocarbon, provide some of our best information on the rates of accumulation and loss of organic matter from soils. Even in situations where C loss or gain over decades can be measured directly, such as in long-term cultivated plots, isotopes give important clues as to what kind of C is lost and how long it might take to accumulate.

Measures of C dynamics using stable C isotopes rely on differences in the fractionation of ^{13}C and ^{12}C during the fixation of CO_2 during photosynthesis. For example, in tropical pastures where plants with predominantly C3 photosynthetic pathways are replaced with C4 grasses, the large difference in the isotopic signature of C can be used to distinguish the rate of loss of forest soil carbon from the rate of accumulation of grass-derived C. Changes in ^{13}C have proved especially useful in studying tropical soils that have been converted from C3 forests to pastures dominated by C4 grasses; see reviews by *Balesdent et al*. [1987, 1998, 2000]; *Bernoux et al*. [1998]; *Volkoff and Cerri* [1987]. However, there are a number of potential complications in applying the stable isotope methods, including uncertainties in the stable isotope signature of the end-member vegetation sources [*Veldkamp and Weitz*, 1994]. Degraded pastures often include not just C4 grasses but a mix of shrubs and grasses which makes identification of an "end-member" isotopic signature difficult. Plant roots and leaves, and the different tissues present within them, may have different isotopic signatures and contribute disproportionately to stabilized soil organic matter pools. Further, many models interpreting changes in stable isotopes assume that the turnover time of forest C3 and grass C4 inputs is the same, which is not necessarily true given the different tissue chemistry for these plant types [*Wynn and Bird*, 2007]. Stable isotopes cannot be used to assess changes in C dynamics compared to intact forests experiencing no vegetation change.

Radiocarbon may be used to study C dynamics on two timescales. Prior to 1950, the radiocarbon age can be used to infer the dynamics of C cycling on century to millennial timescales [e.g., *Paul et al.*, 1997]. Atmospheric testing of nuclear weapons (which largely took place between 1960 and 1964) nearly doubled the amount of ^{14}C in atmospheric CO_2 and produced a global isotopic tracer for organic matter dynamics [*Trumbore*, 2000, 2006]. As with stable isotopes, there are complications to the interpretation of radiocarbon data in terms of the dynamics of the carbon in soils. First, radiocarbon provides a measure of the time elapsed since the C in organic matter was first fixed from the atmosphere by plants; it thus includes the time spent in living plant tissues in the estimate of ^{14}C "age." For example, wood detritus from a century-old tree might decompose within a few decades, but the "age" of the decomposing C will integrate both timescales; any attempt to infer decomposition rates from radiocarbon must account for this effect. Also, unless multiple samples are available from various points in time since 1950 [e.g., *Telles et al.*, 2003], several different models of C dynamics can be used to explain the same set of observations of radiocarbon content. Interpretation of radiocarbon data should report the sensitivity of model-derived turnover times to such uncertainties.

Changes in the ^{14}C of Amazonian soil organic matter since 1960, combined with observations of the rate of change of soil C stocks and ^{13}C signatures in disturbed soils, show definitively that C in soils has several intrinsic timescales of accumulation and decomposition and that modeling all soil organic matter as a homogeneous pool with a single turnover time is clearly overestimating response on decadal to century timescales [*Telles et al.*, 2003]. For example, initially rapid changes in the amount and ^{13}C signatures of organic carbon in surface soils following conversion to C4 grass-dominated pasture demonstrate the presence of fast-cycling organic matter pools. However, the persistence of SOM that is hundreds to thousands of years old and derived from C3 plant sources even in decade-old pastures [*Tiessen et al.*, 1994; *Camargo et al.*, 1999] signifies that a large fraction of SOM cycles much more slowly.

Attempts to physically or chemically separate SOM into fractions that cycle on intrinsically different timescales have met with limited success. Nonetheless, some generalizations can be made. Turnover times are fastest for low-density (<2 g cm^{-3}) or sand-sized (>63 μm) organic matter that mostly represents relatively fresh litter and root detritus [*Lehmann et al.*, 2001; *Paul et al.*, 2008], while the oldest C in soils is strongly associated with clay mineral surfaces [*Telles et al.*, 2003]. While sources of low-density organic matter (e.g., dead roots) do not show large trends with soil depth [*Trumbore et al.*, 2006], other fractions do increase in age with depth (Figures 1 and 2).

In forest soils, *Telles et al.* [2003] identified three different components of soil C that cycled on different timescales: (1) light (density <2 g cm^{-3}) fraction, particulate organic material that could still be identified as plant detritus with more depleted $\delta^{13}C$ signatures, and radiocarbon ages of decades or less; (2) mineral-associated C that is solubilized in acids and bases, ^{13}C-enriched, with radiocarbon ages of decades at the surface to millennia at depths >20 cm, and (3) nonhydrolyzable,

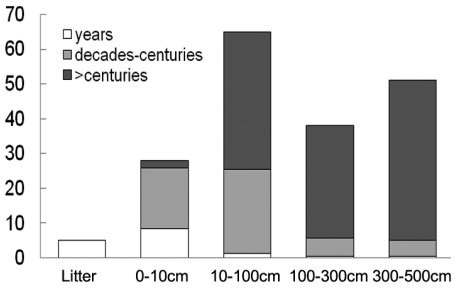

Figure 1. Age distribution of carbon stored in soils with depth. Data are from *Telles et al.* [2003] and *Trumbore et al.* [1995]. Ages of C are derived from radiocarbon data in soil organic matter fractions and inferred from modeling C fluxes and isotopic signatures.

^{13}C-depleted C strongly associated with clay surfaces with ages of hundreds of years (in the 0- to 10-m layer) to >20,000 years (deeper than 20 cm). The amount of C in each of these fractions is shown in Figure 1. Soil texture, in this case clay content, exerts a major control on the amount of slowly cycling carbon and therefore influences the storage and dynamics of carbon in tropical forest soils. *Telles et al.* [2003] also demonstrated predictable relationships between the ^{13}C and ^{14}C content of soil organic matter and soil clay content that are potentially useful for scaling relationships among Oxisol and Ultisols with similar soil age and vegetation.

As shown in Figure 2, of the total ~100 Mg C ha^{-1} in the top meter of an Oxisol in an intact forest, ~5% is in forms with ages of several years or less, ~28% in forms fixed from the atmosphere decades-centuries ago (the half of this in the upper few centimeters is in the form of decades, the remainder centuries), and the rest (>65%) in forms with ages averaging many thousands of years. By contrast, the mean "turnover time," one would calculate for soil C to 20–30 cm depth, defined as the inventory of carbon divided by the rates of C addition or loss would be ~40 Mg C ha^{-1}/~7 Mg C ha^{-1} a^{-1} or ~6 years (Tables 1 and 2). Increasing the depth interval to 100 cm would more than double the C inventory without increasing C inputs significantly (~100 Mg C ha^{-1}/~7 Mg C ha^{-1} a^{-1} or ~14 years). The mean radiocarbon ages of carbon for the

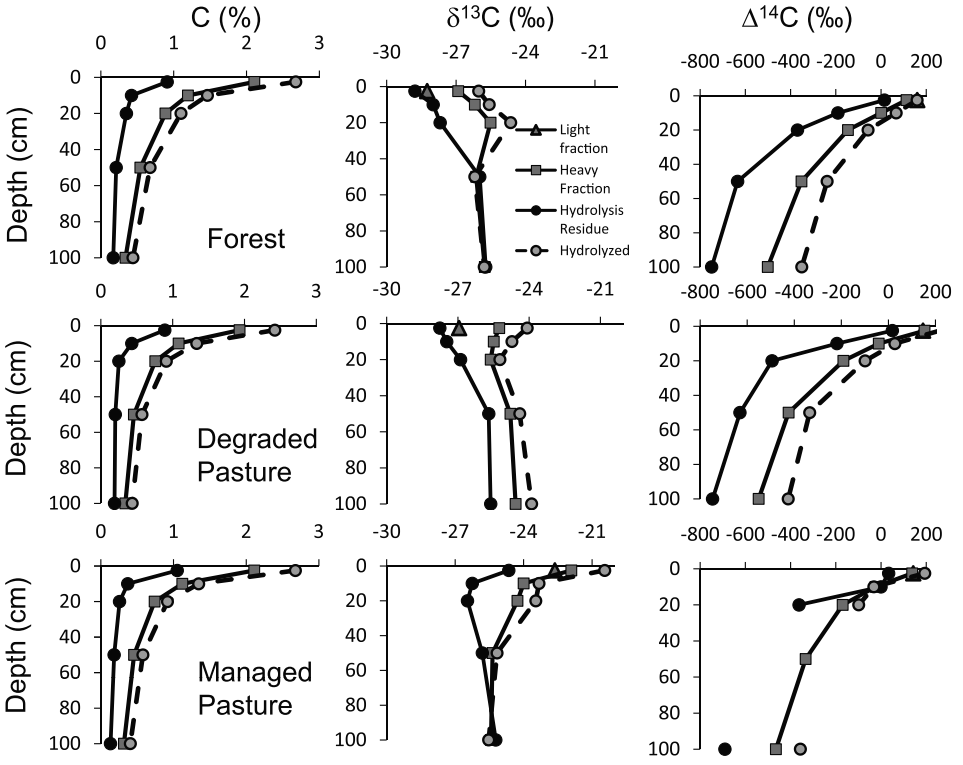

Figure 2. Comparison of C in soil fractions versus depth for three ecosystem types: forest, degraded pasture (~20 years after conversion) and managed (fertilized, planted with productive grass) pasture developed ~5 years previously on degraded pasture. Data are from sites near Paragominas, PA and reported by *Camargo et al*. [1999] and *Telles et al*. [2003]. Note low ^{14}C content of the hydrolysis residue (dark solid line) compared to hydrolyzed carbon (dashed line) and changes in ^{13}C for all fractions in pasture management near the surface.

upper 20 cm of soil range from 200–380 years for clay-rich soils. All of these measures show that C cycles on different intrinsic timescales; the mean "turnover time" averages over components with much younger and older ages. The use of the mean "turnover time" would overestimate short-term (i.e., decadal) responses to disturbance [*Telles et al.*, 2003].

The above results are for well-drained upland soils with high clay content. Radiocarbon data are rare for seasonally flooded Spodosols or wetlands. *Telles et al*. [2003] reported data for a Spodosol near Manaus that showed that the majority of the C to 40-m depth was fixed in the post-1963 period. Given the hydromorphic nature of this material, it is likely that these soils are not representing steady state conditions, more work is needed to understand the dynamics of these soils.

5. RATES OF SOIL C CHANGE WITH LAND MANAGEMENT

Forest clearing for pasture or agricultural use results in changes in the soil physical, biological, and chemical environment. Soil physical structure is altered, along with soil temperature and moisture regimes. Allocation patterns above- and belowground are altered, and thus so are the quality, quantity, and vertical distribution of litter inputs to soil. Changes in C stocks are largest and most rapid in sand-sized organic matter (equivalent to low-density or particulate organic matter), followed by silt-sized fractions, with little or no change in the organic material associated with clay-sized particles [*Shang and Tiessen*, 1997; *Lehmann et al.*, 2001].

Figure 2 shows typical differences in fractionated soil C and isotopes between intact forest (top panel), a 17-year-old degraded pasture (middle panels), and a managed pasture (lower panel). The fertilized and managed pasture has gained carbon, and the degraded pasture lost carbon, compared to the forest soil. These C gains or losses are pronounced only in the surface ~20 cm and are accompanied by changes in ^{13}C associated with the C4 pasture grasses. Largest changes in ^{13}C and C content are observed in hydrolyzable components (organic matter associated with sesquioxides and weakly bound to clay minerals), compared to nonhydrolyzable residues. Although changes in C with depth are attenuated and difficult to detect with depth, changes in rooting depth and root production rates between forest and pasture grasses can be accompanied by significant additional C gains above ~2 m (for very productive grasses) and losses below ~2–3 m [*Trumbore et al.*, 1995; *Camargo et al.*, 1999]. These changes will occur over decades, and there may be a delay associated with root lifetimes that can range up to a decade or more [*Trumbore et al.*, 2006]. Similarly, models of C increase may have to take into account the time lag required for shorter-lived grass roots to increase inputs of dead root material to the upper few meters of soil; this may explain why *Camargo et al.* [1999] predicted larger-than-observed changes in the ^{13}C of SOM in their managed pasture site.

On conversion from forest to pasture, C inventory may increase, decrease, or stay about the same [*Shang and Tiessen*, 1997; *Neill and Davidson*, 1999; *Holmes et al.*, 2006]. Summaries of detailed chronosequence studies [*Neill and Davidson*, 1999] and more spatially extensive data sets [*Holmes et al.*, 2006] demonstrate that the overall direction and magnitude of change in SOC following forest clearing is broadly predictable from the original forest soil carbon content and pH. Decreases in C occur when initial forest C content is high and replaced by less productive pasture or agricultural vegetation. Increases in C occur when initial forest surface soil C content is low, and vegetation is replaced with highly managed and fertilized productive grasses. Rates of initial change in C stocks can be rapid (~5% per year) but decline over time. For sites with declining C content, reductions in the amount of fast-cycling C mean overall reduction in the supply of soil nutrients derived from mineralization of organic matter and loss of fertility without additional fertilization [*Tissen et al.*, 1994].

Using an ecosystem C model (Century) where the turnover time of C in active slow and passive pools is controlled by litter quality, climate, and soil texture, *Schimel et al.* [1994] showed the predicted average residence time for SOM in the upper 20 cm of soils in tropical regions can range from <14 years up to several decades. However, data in Figure 2 suggest that the model underestimates the residence time of C in passive C pools (the inventory-weighted age of C for the soil in Figure 2 is several hundred years). Such differences between models and observations affecting very long timescales may not be important for predictions of changes on timescales of a few decades. For example, *Cerri et al*. [2007a] have used the Century model to successfully reproduce observed changes in C inventory and ^{13}C in several pasture sites following conversion in Rondônia.

Studies of soil C change with land use at specific sites are useful for understanding the dynamics and local controls on C cycling. However, spatial extrapolations require assumptions about the areal extent of land management practices. For the state of Rondonia, *Holmes et al.* [2006] were able to take advantage of a large data set of C large regional soil profile database and remotely sensed classification of land cover to estimate the net C storage or loss associated with land cover changes across a range of original soil properties and management regimes. Although individual locations could have either large C gains or losses, the spatially averaged result was a net small loss of C for the state of Rondonia. Hence, while site data are important for managing individual

plots for soil fertility, uncertainties in net C storage or loss at the landscape scale may depend more on the distribution of management across the landscape.

Cerri et al. [2007b] and *Falloon et al.* [2007] have predicted future soil C stock changes based on scenarios of land cover and climate change for the Amazon Basin. *Cerri et al.* [2007b] predict a ~7% decline in Amazon soil C stocks (0- to 30-m depth) between 2000 and 2030, balancing losses in newly cleared areas with increases in areas abandoned or used for secondary forest regrowth. *Falloon et al.* [2007] predicted very large net loss of C with climate change alone, due to drying and decrease of forest C productivity. However, the magnitude of their estimated loss (a decline from 45 Pg C to 23 Pg C between 2000 and 2100) is too large given that >70% of the soil C in the 0- to 30-cm layer is in soil organic matter fractions with turnover times longer than centuries. The large predicted losses likely result from the use of a single pool model for soil C that overpredicts C changes in the short term [*Knorr et al.*, 2005]. Using more realistic models of C dynamics, losses from soils will be small to minimal compared to changes in aboveground biomass that occur with deforestation and reforestation. It is the fate of tree biomass that will determine the overall magnitude of tropical land use change as a source of C to the atmosphere over the next decades.

5.1. Anthropogenic Soils

Soils with the greatest C storage in the Amazon are the "terra preta do Indio" or Indian Black Earth soils. These soils, known as Antrhoposols, or human-generated soils, are associated with areas inhabited by indigenous peoples from 500 to 2500 years ago and abandoned after European arrival [*Woods*, 2003]. Organic C is enriched to 1 to 2 m depth compared to adjacent soils, with much of the stabilized C thought to be in the form of charred or "black" C; [*Glaser et al.*, 2001]. The carbon stabilized in these soils remains hundreds to thousands of years after they were abandoned. Indian Black Earths are also high in phosporous, CEC, pH, and base saturation and, consequently, are more fertile than surrounding soils. For a review, including the potential for people to manage soils to sequester carbon, in essence to recreate these soils, see *Lehmann et al*. [2003].

6. TIME LAGS BETWEEN PHOTOSYNTHESIS AND RESPIRATION AND THE ESTIMATION OF C SINK POTENTIAL IN INTACT (UNMANAGED) FORESTS

Undisturbed tropical forests have been proposed as a potentially large sink for anthropogenic carbon based on inversions of regional atmospheric CO_2 concentration variations [*Stephens et al.*, 2007]. What role might soils play in a regional C sink? The capacity for an ecosystem component such as soil to serve as a net sink of carbon may be estimated from the magnitude of gross fluxes and the time C resides in each cascading C pool [*Fung et al.*, 1997; *Thompson et al.*, 1996]. Because of the heterogeneous nature of soil carbon, the age of C respired from soils is nearly always younger than the age of bulk C stored in soils [*Trumbore*, 2000]. This is due to the fact that most of the C being respired is from pools that cycle slowly, while pools with the longest residence times have the largest C stocks.

An estimate of the time lag between photosynthesis and respiration based on radiocarbon data [*Trumbore et al.*, 2006; *Telles et al*., 2003], stemwood production [*Vieira et al*., 2005] and respiration fluxes [*Chambers et al.*, 2004] in a central Amazon forest near Manaus is illustrated in Figure 3. Even though much of the C fixed is respired autotrophically and over a relatively short time period, the remaining C either stays in living leaf and root tissues for an average 1–3 years (leaves), or longer for the ~50% of litterfall that is not leaves, and 5–10 years (roots), then decomposes rapidly. Overall even though only ~20% of the total ecosystem respiration is estimated to be from microbial decomposition

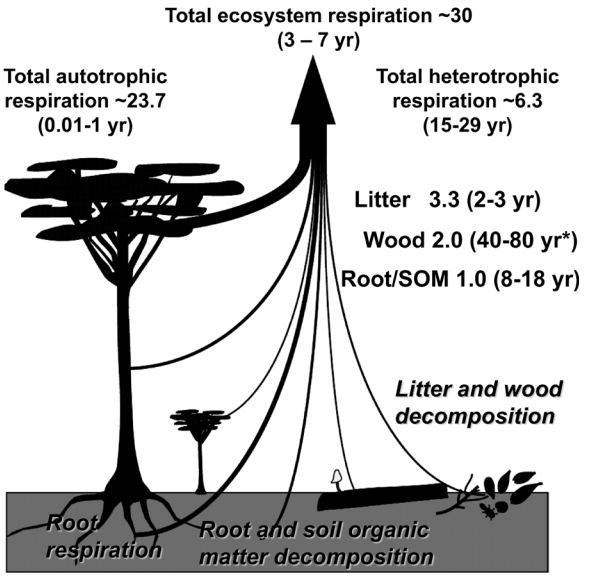

Figure 3. Calculation of the mean age of ecosystem respired CO_2. Fluxes are based on *Chambers et al.* [2004]; mean age of decomposing wood based on *Vieira et al.* [2005] and wood decomposition rates of *Chambers et al.* [2001b]; age of C derived from dead root decomposition from *Trumbore et al.* [2006]; mean age of litter respired CO_2 from *Brando et al.* [2008] and unpublished data of P. B. Camargo and S. Trumbore (2008).

of plant residues and soil organic matter, the time lags involved are significant. Results in Figure 3 generally agree with those published in *Fung et al.* [1997], which estimated the mean age of heterotrophically respired C to be 24 years for broad-leaved evergreen tropical forests.

Models of C dynamics indicate vegetation and soils under conditions of increased forest productivity should remain sinks for as long as inputs outstrip the increases in decomposition or mortality rates. For example, *Chambers et al.* [2001a] used an individually based model of tree growth and mortality, forced with a 25% increase in productivity over a period of 50 years (an increase of 0.25%/year) to estimate maximum C sequestration rates of ~0.5 Mg C ha^{-1} $year^{-1}$ in woody biomass in the period over which forcing was applied. Models of C dynamics in soil organic matter calibrated using observations of ^{14}C [*Telles et al.*, 2003] used the same forcing scenario (assuming all inputs would increase at the same rate as overall NPP with no additional time lag) and showed that rates of net C accumulation in soil (assuming decomposition rates remained unchanged) would not be larger than about 0.1 Mg C ha^{-1} a^{-1} in the upper 40 cm of soil. According to *Trumbore et al.* [1995], this rate could potentially be doubled if changes in root productivity below 40 cm were taken into account. Hence, the overall predicted rate of C sequestration would be in the order of 0.7 Mg C ha^{-1} a^{-1}, with the majority C storage in aboveground biomass.

Given that the increased NPP derived from CO_2 fertilization would not be expected to increase quite as rapidly, a more realistic estimate, using a simple box model that forces increases NPP in a tropical forest (fluxes in Figure 3) with a β factor of 0.2 is represented in Figure 4 [see also *Chambers and Silver*, 2005]. Soil C storage lags vegetation storage and is responsible for a much smaller fraction of the total estimated C sink for the early 2000s (at ~13% of a total ecosystem C sink estimated at ~0.12 Mg C ha^{-1} a^{-1}). However, soil C storage will continue for decades after NPP increases cease because of the time lags in living vegetation.

Multiplying an estimated sink of 0.12 Mg C ha^{-1} a^{-1} times the area of the Brazilian Amazon (~5×10^8 ha) would result in a net C sink associated with CO_2 fertilization of 0.07 Pg C a^{-1} , 13% of which would be in soil. A sink of this magnitude is not large enough to balance regional C losses from deforestation and is not easily detectable given current methods for determining forest C balance. Recent measurements of permanent plots in Panama and Malaysia [*Feely et al.*, 2007] have documented declines in stem wood increment at the stand level, rather than the increases that might be expected with CO_2 fertilization. Long-term analysis of permanent plot data in the Brazilian Atlantic forest have similarly shown a tendency for rapid C loss associated with sudden mortality events, followed by periods of more rapid tree growth

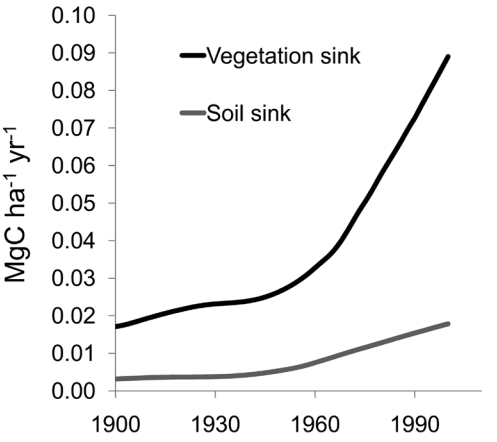

Figure 4. Estimated C sink for the forest ecosystem pictured in Figure 3, assuming a B factor of 0.2 for CO_2 fertilization, the record of measured atmospheric CO2 from 1800 to 1990 and a linearly increasing rate of CO_2 increase from 1990 to 2010 that matches observations through 2007 (CO_2 concentrations in 2010 are estimated at 395 pap. The soil sink lags the vegetation sink and is only ~13% of the total sink of 0.12 Mg C ha^{-1} a^{-1} estimated for the year 2007.

[*Rolim et al.*, 2005]. Even if productivity increases, it will likely be associated with either changes in vegetation/litter quality or allocation among leaves, stems, and roots (particularly involving rooting depth). As such, changes are likely to occur over decades; it is difficult to predict the overall impact on predicted C stores in soils. What we can conclude is that changes in soil C stocks have limited potential to offset current deforestation sources; again it is the more dynamic vegetation C pools that will dominate any response.

Variations in NPP from 1 year to the next are larger than the long-term trends calculated above. For example, interannual variations in litterfall and tree biomass increment of order ~1 Mg C ha^{-1} a^{-1} each have been reported [*Rice et al.*, 2004; *Vieira et al.*, 2006]. Given a deviation of 2 Mg C ha^{-1} a^{-1} (out of ~30) and a time lag of ~3 years between photosynthesis and ecosystem respiration (75% respired in same year, 25% with average of 12 years), the expected interannual variation in net C storage/loss can reach ±0.25 Mg C ha^{-1} a^{-1}. If such variations are spatially coherent variations across the Amazon Basin (e.g., like those associated with ENSO climate anomalies), they have the potential to contribute significantly to observed interannual variations in global CO_2 accumulation in the atmosphere.

7. OUTSTANDING QUESTIONS

The major conclusion of this review is that changes in soil C in response to land use or climate change, or even CO_2 fertilization will be minimal compared to changes in

aboveground C pools over the next century. Why then argue for more research in Amazonian soil C? Most importantly, soil organic matter largely determines the fertility and sustainability of pastures and agricultural land [*Tiessen et al.*, 1994]. More studies are needed to elucidate the processes stabilizing C in soils and links between how C is stabilized and how long it remains in the soil. Cycling of organic matter is inexorably linked to biogeochemical cycling of water, nitrogen, phosphorous, and other elements, which are keys to soil fertility and agricultural sustainability. Research linking C and nutrient dynamics is clearly needed.

While the overall dynamics of C in the major soil orders found in the Amazon basin have been explored in a preliminary way, the amount of data, especially for intact forests, remains small. Recent studies have demonstrated some major surprises in how intact forests cycle carbon, especially: (1) the low carbon use efficiency of these forests [*Chambers et al.*, 2004] and (2) the longevity of live fine root biomass [*Trumbore et al*., 2006]. However, these conclusions are based on data from very few sites and may not be representative of the Amazon basin as a whole. In particular studies of C stocks and fluxes in wetland and seasonally flooded soils need to be augmented with isotopic measures of C dynamics on decadal and longer timescales.

Some of the C in soils is very old (>25,000 years), and retains evidence of C4 vegetation sources likely dating from the last glacial period [*Saniotti et al*., 2002]. Little is understood of the processes that can store C for so long—and whether it is in the form of char ("black C") or other forms. Amazon landscapes changed dramatically over the last glacial cycle, and some of the properties we observe in soils today may reflect conditions at that time.

Another area where more research is needed is in the tools for extrapolating soil C stocks and dynamics from point measurements to landscapes. *Holmes et al.* [2006] demonstrate the importance of spatial approaches in determining what the important factors for assessing C at different spatial scales are. It is particularly important to build an understanding of the key processes involved in stabilizing and destabilizing carbon, in particular, the role of minerals, soil fauna, and aggregates, and how these may vary regionally and in response to land cover change.

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